1. IC5.3: The Effect of Stability on the Response to Internal Forcing in the Atmosphere

Instructor Notes: Welcome to the winter AWOC IC 5 on precipitation forcing. This is Lesson 3 – The Effects of Stability on the Response to Internal Forcing in the Atmosphere. This lesson is presented to you by Phil Schumacher and David M Schultz. This lesson is 30 slides and should take you 30 minutes to complete.

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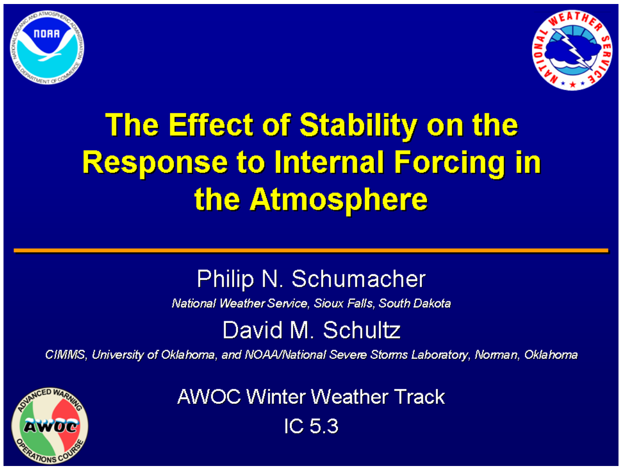

2. Learning Objectives

Instructor Notes: This lesson has four learning objectives. First, you should be able to define static, intertial, and symmetric stabilities. Next, you should be able to describe the processes that change static and symmetric stabilities. Third, you should be able to identify two advantages and one disadvantage to using EPV instead of M and saturated equivalent potential temperature surfaces for diagnosing symmetric instability. Lastly, describe how the shape and intensity of the frontal circulation varies with symmetric stability.

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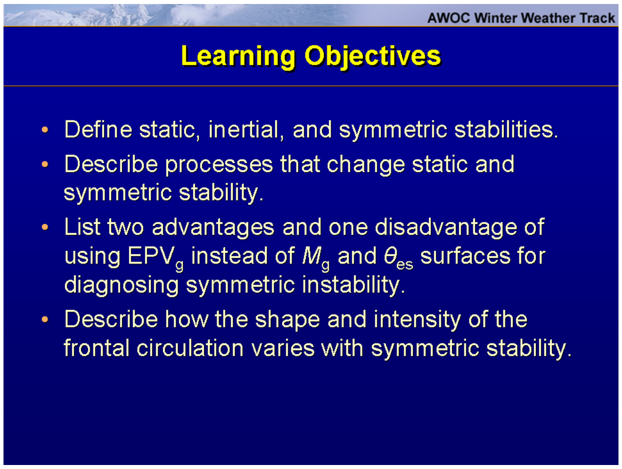

3. Performance Objectives

Instructor Notes: There are two performance objectives for this lesson. You should be able to determine what layer to examine EPV when forecasting. Secondly, you should be able to determine the impact that stability can have on snowfall rate and the width of a snowband.

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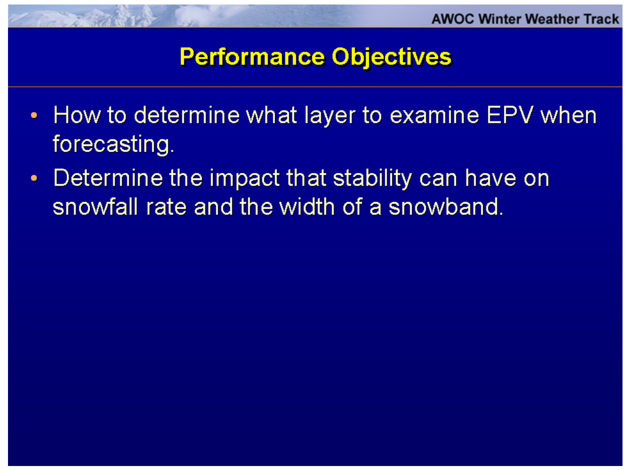

4. The Atmospheric Response to Frontogenesis Depends Upon the Strength of the Frontogenesis and the Stability

Instructor Notes: In Lesson 2, we learned about frontogenesis. When frontogenesis occurs, a secondary ageostrophic circulation results that attempts to restore thermal wind balance. Specifically, in the presence of frontogenesis, a thermally direct circulation results. The strength and depth of a secondary ageostrophic circulation is dependent upon the strength of the frontogenetical forcing and on the stability. In Lesson 2, we saw that all things being equal, stronger frontogenesis leads to greater vertical motion. Later in this presentation, we will be more precise in what we mean by stability. In the meantime, let’s consider two types of stability: static stability and inertial stability. Static stability is a measure of the resistance of the atmosphere to vertical displacements. Inertial stability is a measure of the resistance of the atmosphere to horizontal displacements.

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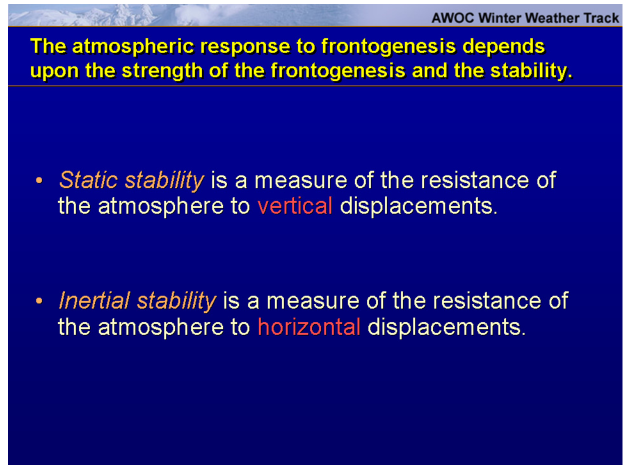

5. Perturbations Placed in an Unstable Environment Grow Larger

Instructor Notes: The definition of an unstable environment is one in which perturbations placed in that environment grow larger. It doesn't matter what kind of environment we're talking about. A pencil standing on its pointed end on a table is unstable. Any perturbations to the pencil will result in the perturbation growing and the pencil falling over. The atmosphere can be unstable to vertical or horizontal displacements. One type of instability to vertical displacements is called static instability. You should be familiar with the concepts of static stability. The condition for static instability is that the environmental lapse rate is between the dry and the moist adiabatic lapse rate, or equivalently the saturated equivalent potential temperature decreases with height. The saturated equivalent potential temperature is the equivalent potential temperature an air parcel would have if it were brought to saturation at the same temperature and pressure by evaporating water into the parcel. Conditionally unstable environments may eventually undergo vertical motions in the form of deep moist convection that overturn the unstable layer and release the instability. You may be less familiar with inertial instability. Inertial instability is not common in the free atmosphere, but may be found on the anticyclonic-shear side of jets where the absolute geostrophic vorticity is less than zero. Inertially unstable environments can develop a form of "horizontal" convection in which ageostrophic circulations develop to restore balance. In this regard, the release of inertial instability is analogous to the release of static instability to form deep moist convection. Thus, for static stability, the greater the saturated equivalent potential temperature decreases with height, the more statically unstable the air is. The smaller the absolute geostrophic vorticity is, the lower the inertial stability.

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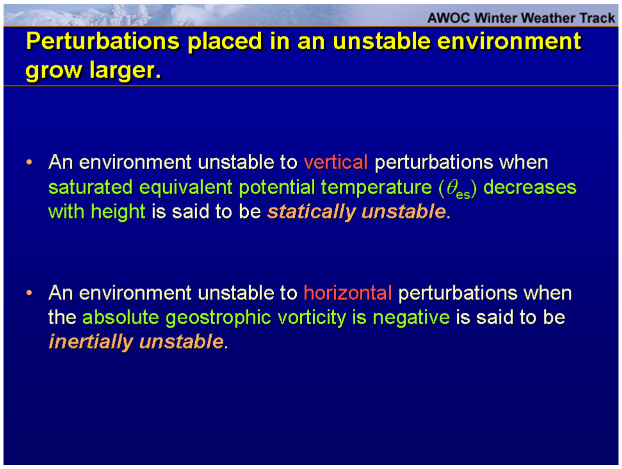

6. The Rate of Adiabatic Cooling Depends on the Static Stability of the Environment

Instructor Notes: For example, let's consider how the static stability can affect the vertical response to frontogenesis. One can define stability as the difference between the environmental and the dry (or moist) adiabatic lapse rate. For a given amount of forcing (i.e. frontogenesis), when stability is large [a large difference between the environmental lapse rate and dry (moist) adiabatic lapse rate] then the vertical motion is most likely going to be small. On the other hand, when stability is small, then there will be a lot of vertical motion. Also, note that whether a parcel is saturated or not can have a large effect on the stability since the moist adiabatic lapse rate is smaller than the dry adiabatic lapse rate (9.8 degrees C/km vs. ~6.5 degrees C/km). What can be a stable environment when the atmosphere is unsaturated, can become much less stable when the atmosphere becomes saturated. This would mean a large increase in vertical motion once the atmosphere is saturated.

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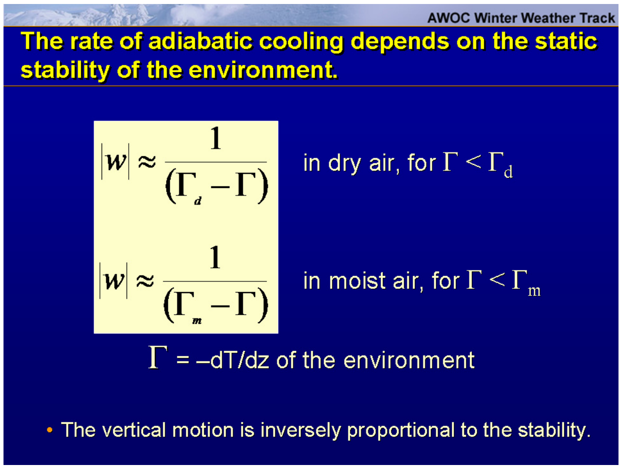

7. Stability Impacts the Degree of Coupling Between the Upper- and Lower-Level Circulations

Instructor Notes: Recall from Lesson 1 that potential vorticity anomalies will induce winds at levels far removed from where the anomaly is located. The Rossby depth determines how deep into the atmosphere the influence of a particular wave will be. While the horizontal extent of the wave is one factor, stability plays a large role in determining the ability of an upper level PV anomaly to “communicate” with a low-level front. This relationship is best illustrated by comparing summer and winter. In winter, strong PV anomalies that move over cold anticyclones will result in little vertical motion or surface response because of the high stability within the air mass. In summer, smaller and weaker waves can produce a large response because the stability is generally near moist adiabatic. Therefore, knowledge of the stability between a low-level front and upper-level wave is critical to understanding the strength of a coupled jet/front circulation.

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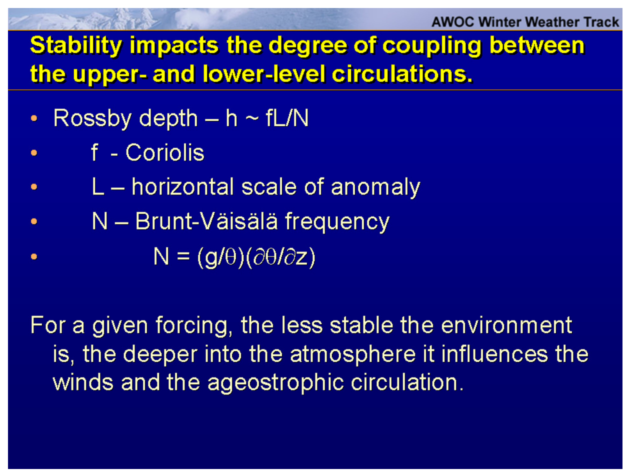

8. Case 1

Instructor Notes: As we saw from the relationship between stability and vertical motion, high static stability means weak vertical motion. So, the vertical component of the circulation will be shallow and weak. At the same time, low inertial stability means that air parcels will accelerate far away from their initial horizontal position. Thus, the horizontal component of the circulation will be large and strong. Therefore, we would see a broad but shallow circulation in the event of high stability and low inertial stability.

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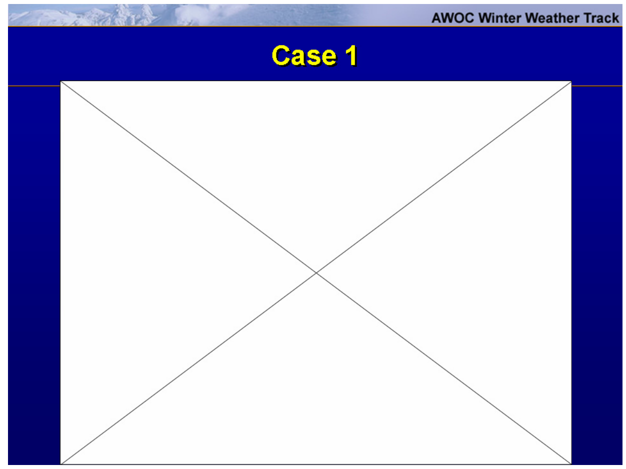

9. Case 2

Instructor Notes: The other extreme would be weak static stability and high inertial stability. With low static stability, the vertical component of the secondary circulation will be large and deep. With high inertial stability, the horizontal component of the secondary circulation will be small. Thus, the circulation is narrow in the horizontal but extends much more deeply in the vertical.

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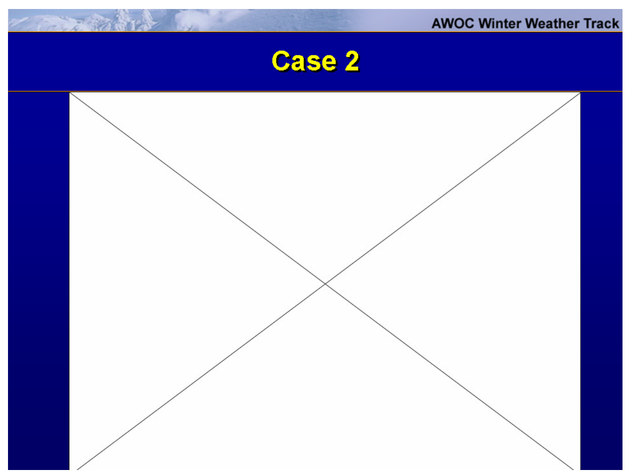

10. Symmetric Instability is a Generalization of Static and Inertial Instability

Instructor Notes: Instability in the atmosphere can be viewed more generally. Specifically, the instability resulting from vertical displacements and the instability resulting from horizontal displacements can be generalized into instability resulting from slantwise displacements. Symmetric stability is a measure of the resistance to slantwise ascent by parcels. The atmosphere can be symmetrically unstable but inertially and statically stable. Symmetric stability plays a large role in determining the strength and width of the ageostrophic frontal circulation. When there is small symmetric stability, the result will be an intense and narrow updraft. When symmetric stability is large, a broad and weak updraft.

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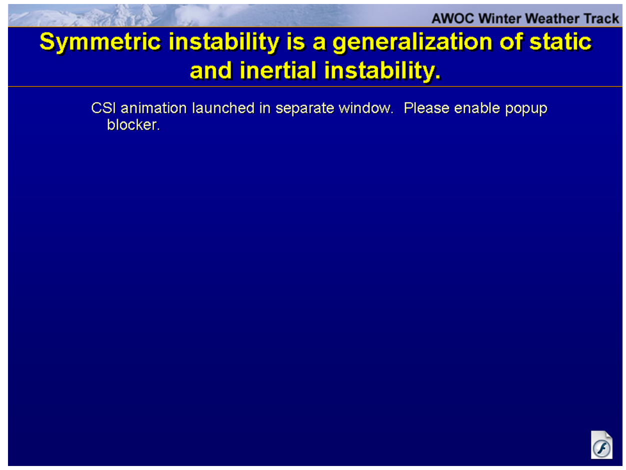

11. The Symmetric Stability Affects the Vertical Motion

Instructor Notes: Thorpe and Emanuel (1985) ran two simulations of a 2-D semi-geostrophic model with same frontogenetic forcing. The only difference they made was to change the symmetric stability in the warm air (through the moist or equivalent potential vorticity, to be discussed later). In the first run, the entire domain had large, dry potential vorticity. We will call this the dry case. In the second run, the potential vorticity was set to near zero on the warm side of the front. This is equivalent to the weakly stable case. The result in the vertical motion is dramatic. While both areas of ascent are slantwise into the cold air, the dry case shows very broad and weak vertical motion. In the "moist" case, the ascent is unbalanced with very strong and narrow ascent on the warm side of the boundary. While the subsidence on the cold side of the front is stronger than the dry case, the higher stability behind the front still means rather broad area of subsidence.

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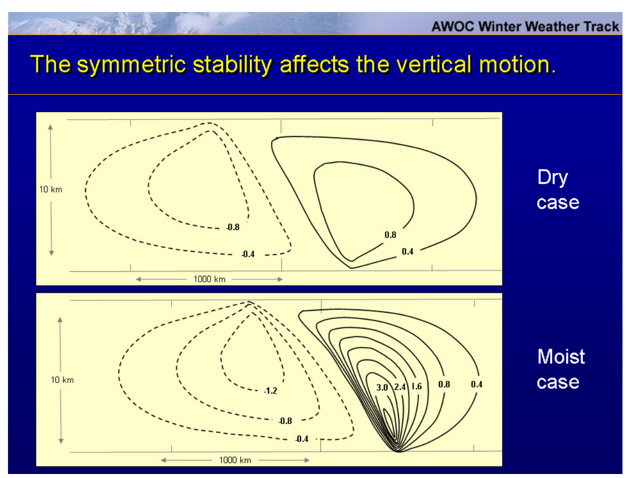

12. Geostrophic EPV (EPV_g) Measures the Symmetric Stability

Instructor Notes: We can calculate the stability of parcels to moist slantwise ascent by calculating Equivalent Potential Vorticity (EPV_g). The equation is similar to dry potential vorticity (PV) except that the full three-dimensional geostrophic vorticity is used, and theta-E (or theta-es) is used instead of potential temperature. The EPV_g equation above has 3 terms. Terms 1 and 2 are vertical wind shear and horizontal temperature gradient, respectively. Near a strong front the horizontal temperature gradient will be large as well the vertical wind shear. Both terms combined will usually be negative (assuming colder air to the north and west in the Northern Hemisphere). Term 3 is absolute geostrophic vorticity and static stability. This term is positive when the atmosphere is inertially and statically stable. There are three conditions when EPV_g is negative: inertial instability, potential (convectively) instability, and symmetric (CONDITIONAL) instability. While inertial instability is rare, to be potentially or conditionally unstable is relatively common, especially in summer. In these cases, the atmosphere is susceptible to upright convection and not slantwise ascent. However, in cases with weak inertial stability and weak upright stability, then the first two terms can force EPV_g to be negative. These are the situations where there can be strong slantwise ascent and even slantwise convection.

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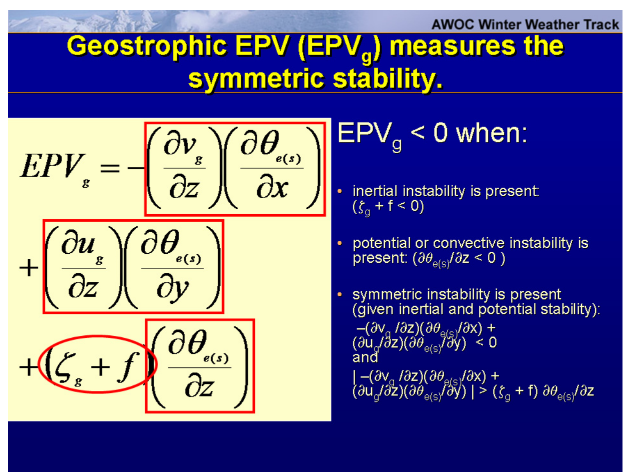

13. Using the Geostrophic Wind or Real Wind in Assessing Symmetric Stability Can Be Important

Instructor Notes: Strictly speaking, evaluating the symmetric stability requires use of the geostrophic wind in the calculation of EPV_g. In high-resolution numerical model output, the geopotential height (and by extension the geostrophic wind) may be noisy, making determining the symmetric stability difficult. In this case, using the total wind may be preferable. In cases where the horizontal and vertical gradients of the geostrophic wind and total wind are similar, the EPV calculated either way will be nearly the same. When large differences occur, the EPV calculated from the geostrophic wind may be found to be more symmetrically unstable.

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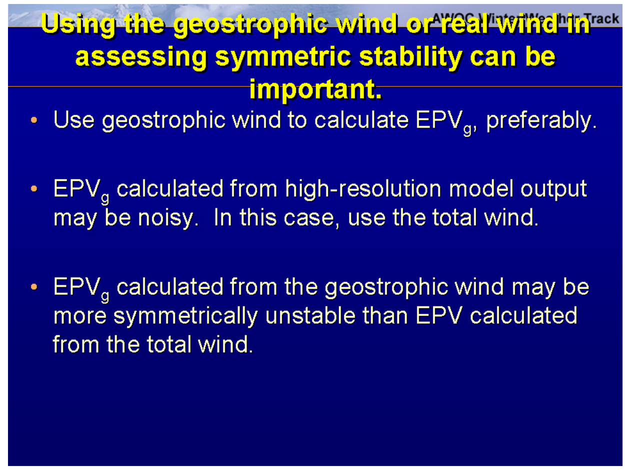

14. Choice of Thermodynamic Variable (θes or θe) is Important

Instructor Notes: As noted in the previous slide, EPV_g can be calculated using θ_e or θ_{es}. The figures show that there can be large differences between the two calculations for the same case. On the left, saturated theta-e is used and, for a large portion of the cross-section above the frontal inversion, saturated theta–e decreases with height, implying convective instability. Therefore the atmosphere would be susceptible to upright convection assuming a saturated atmosphere. On the right, theta–e increases with height through the entire cross-section, implying potential stability throughout. Using saturated theta-e or theta–e when calculating EPV_g can have affect how one would interpret the atmospheric response. Using saturated theta-e, locations where there is conditional instability would be negative and imply the possibility of upright convection. For the cross-section on the right, the entire area is potentially stable. So if one used theta-e to calculate EPV_g, then negative EPV_g would imply potential symmetric instability (PSI) along the entire cross-section. No study has been done to determine which form of EPV is operationally better so forecasters need to be aware of the differences especially if the atmosphere is not saturated.

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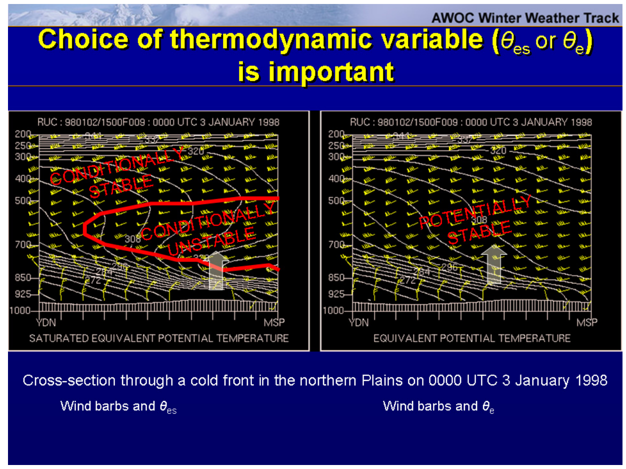

15. Interactive Quiz #1

Instructor Notes: Take a few moments to complete this quiz.

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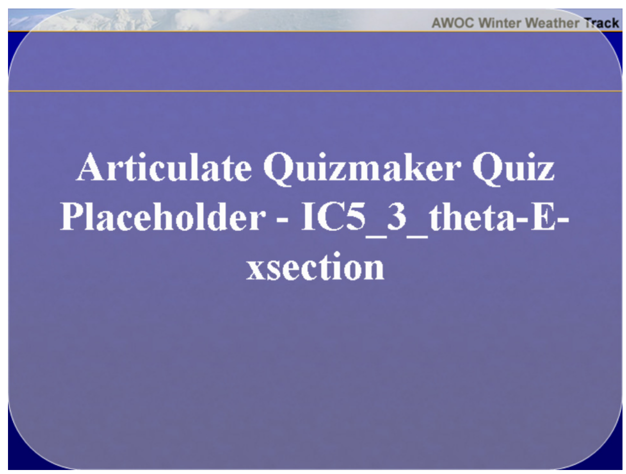
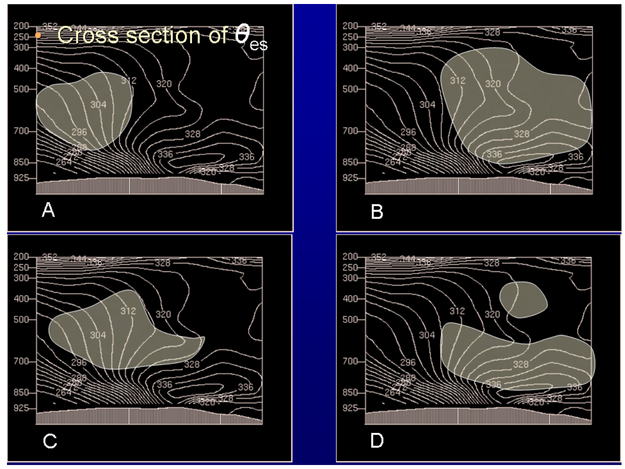


16. Interactive Quiz #1 Review

Instructor Notes: The best answer to the question is the analysis in D where q_e decreases with height. The shaded area in all other choices include some region where q_e actually increases with height.

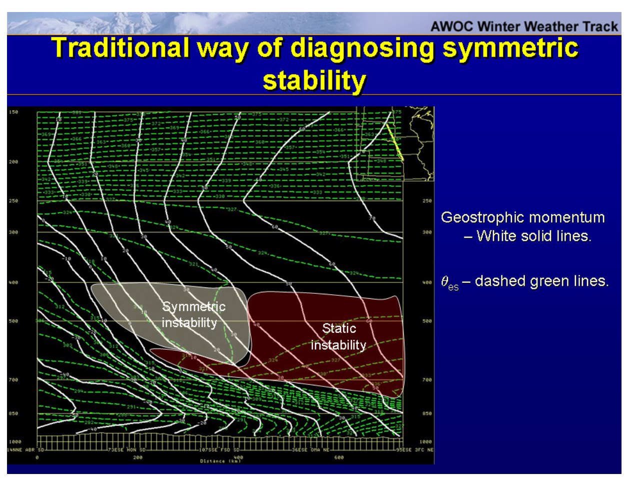
Student Notes:



17. Traditional Way of Diagnosing Symmetric Stability

Instructor Notes: We will compare the M_g–q_{es} technique for identifying areas with symmetric stability to that using EPV. The above figure shows the M_g–q_{es} surfaces. The traditional way that symmetric instability has been taught was to compare the slope of Mg surfaces to those of qes surfaces in a cross section constructed perpendicular to the thermal wind (called the M_g–q_{es} relationship). If the slope of the M_g surfaces is less than that of the q_{es} surfaces, then symmetric instability is present. There are two disadvantages of this approach. First, cross-sections can only tell part of the story. Forecasters need to know the horizontal distribution of instability in order to determine how the frontal band will evolve over. The duration of frontogenesis and low stability over an area is critical for knowledge of how much snow may fall over a location. Second, the validity of the M_g–q_{es} relationship requires the flow is not curved since M is calculated only from the normal component of the wind. Significant direction shear will result in large changes in M that are not the result of speed shear but changes in direction.

Student Notes:



18. Identifying Areas with Symmetric Instability

Instructor Notes: Fortunately, there is a better approach. Negative saturated equivalent geostrophic potential vorticity (MPV_g*) is equivalent to the M_g–q_{es} relationship and does not face those above limitations. Unfortunately, distinguishing conditional symmetric instability from conditional instability is not possible without examining the vertical profile of q_{es} to confirm that it decreases with height. A useful diagnostic approach is to overlay frontogenesis, MPV_g*, and RH in a horizontal map or in a cross-section. Above is the same cross-section as for the M-θ_{es} cross-section above. Notice that the layer which is symmetrically unstably (or weakly stable) is obvious in this cross-section when an image is used to highlight EPV near or below zero. By overlaying θ_{es} we can also easily identify regions of conditional instability and symmetric instability. Notices that regions of conditional instability generally have much lower values (more negative) of EPV than areas that are conditionally unstable. While not shown in this case, one may also want to overlay relative humidity. As with conditional instability, only when parcels are saturated will the instability be realized. As we will see later, while the EPV is very negative on the right (south) side of the cross-section, it is also very dry and coincident with the dry slot. Therefore, this instability will not result in enhanced vertical motion until parcels reach saturation.

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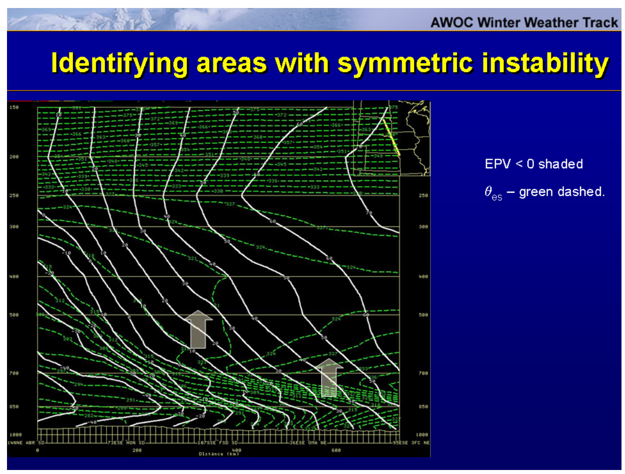
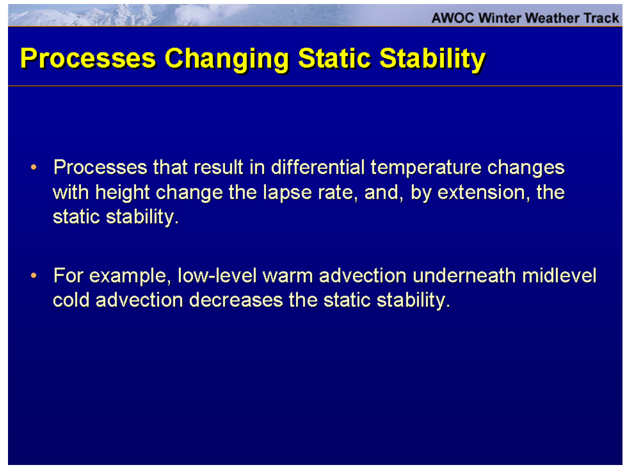


19. Processes Changing Static Stability

Instructor Notes: Situations that are statically stable can be made statically unstable by changing the lapse rate, or producing differential temperature changes with height. For example, one way to produce a different rate of change of temperature between two levels is by having low-level warm advection underneath midlevel cold advection. The combined effect of these two processes will result in a lowering of static stability.

Student Notes:



20. Processes Changing Symmetric Stability

Instructor Notes: Similarly, changes in symmetric stability can be viewed as processes that steepen the θ_{es} contours relative to the M_g surfaces. This will produce a reduction in EPV_g. One location where the θ_{es} contours tend to overturn is the dry slot region of an extratropical cyclone. It is at the leading edge of the dry slot as it rides over the warm front where you can sometimes find heavier precipitation due to the reduction in stability.

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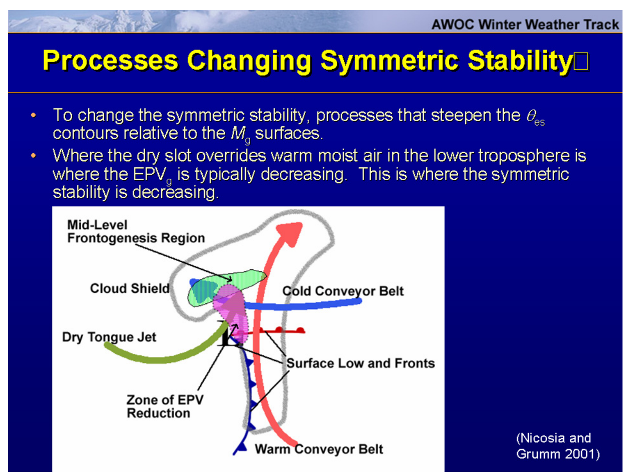
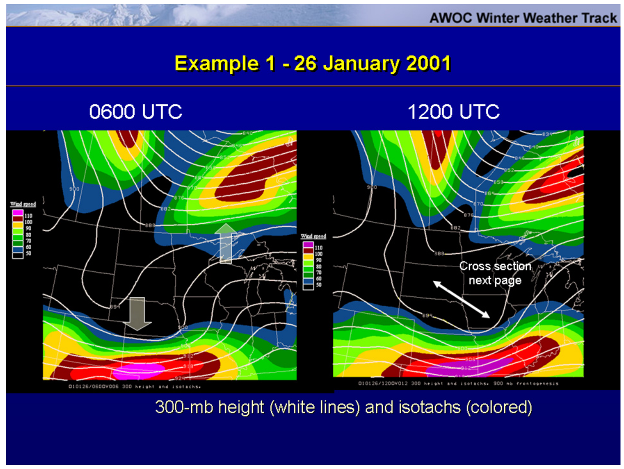


21. Example 1 - 26 January 2001

Instructor Notes: We will now look at two examples which have frontogenesis and an upper-wave but differences in stability. Our first example is from 26 January 2001. The above plots show 300 mb isotachs (shaded) and heights (white lines). This example was a split flow regime with two waves moving across the region: the first moving along the U.S. and Canadian border and the second moving across southern Nebraska and southern Iowa. Both waves are associated with jet streaks where winds were in excess of 100 kts.

Student Notes:



22. Cross-Section of Front and EPV_g

Instructor Notes: The cross-section above shows potential temperature, relative humidity and frontogenesis. This cross-section is taken perpendicular to the low-level front. Notice that the frontogenesis is confined below 850 mb and that above the level frontogenesis is a very stable layer. Were this an unstable environment then the fact that this is a shallow front would not matter: a frontogenetical circulation would develop resulting in precipitation. However, even though it is saturated, the stability is very high above the front. This will limit the strength of the frontal circulation and the ability of the upper wave to couple with the low level front to produce precipitation.

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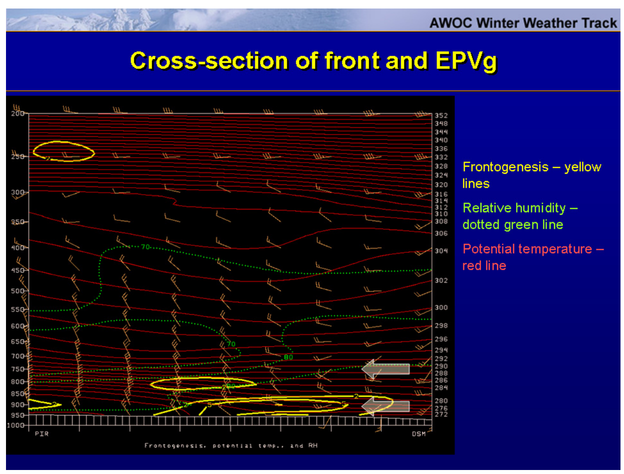

23. 26 January 2001 EPV Plan View Plots

Instructor Notes: These figures shows frontogenesis (thin white lines) and 700 to 800 mb EPV (shaded). What we saw in the cross-section is confirmed when we calculate EPV. Over southeast South Dakota, an area of EPV greater than 1 PVU is situated over the frontogenetic region at 900 mb. As we saw earlier, high stability can act to suppress frontal circulations and therefore one would expect little precipitation across eastern

South Dakota into western Iowa. Across northeast Minnesota, where the stability is lower, the potential for more significant precipitation would be higher.

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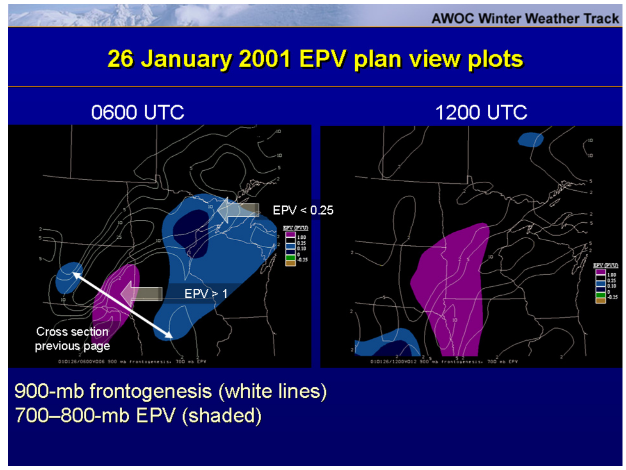
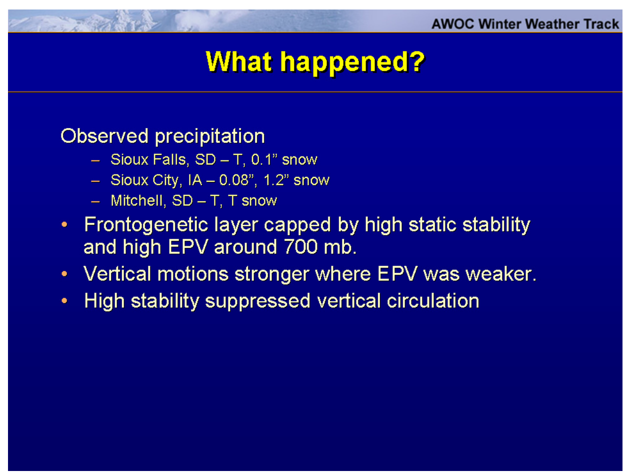


24. What Happened?

Instructor Notes: As a result of the high stability, there was little observed precipitation associated with the front as it moved across. Despite the presence of an upper level wave and low-level frontogenesis, the frontal circulation remained suppressed due to the presence of high stability above the frontal surface.

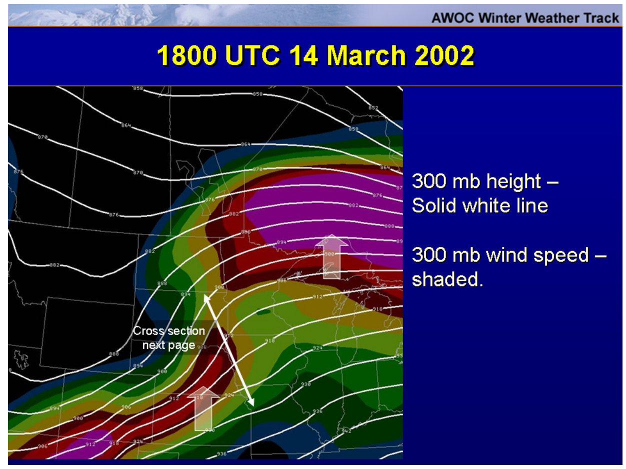
Student Notes:



25. 1800 UTC 14 March 2002

Instructor Notes: The second example is from 14 March 2002 and displayed is 300 mb winds and height. We see two jet streaks. The first is located along the U.S.–Canadian border placing eastern South Dakota and southern Minnesota within the right entrance region of the jet. A second jet streak is located behind an upper-level shortwave moving into Nebraska. From a large-scale perspective the eastern Plains have synoptic-scale support for vertical motion.

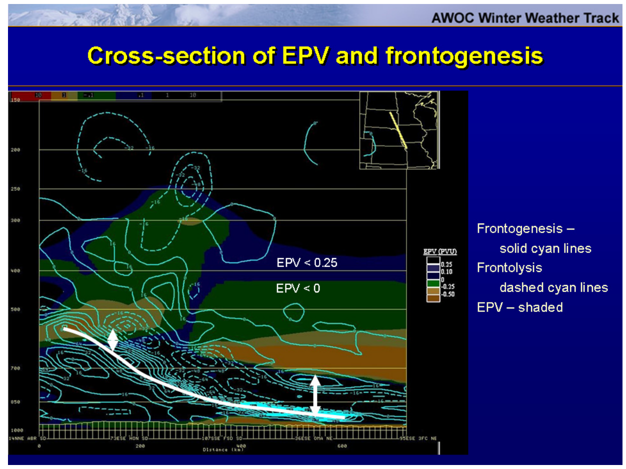
Student Notes:



26. Cross-Section of EPV and Frontogenesis

Instructor Notes: Now let's look at a cross-section perpendicular to the front across eastern South Dakota and western Iowa. In Lesson 2, we discuss how one can examine where the frontogenesis with respect to the Q-vector convergence in a layer near the tropopause. Where the two (nearly) overlay is the level (or layer) where one can get a coupled circulation between the upper-level wave and low-level front. However, there are cases where the upper-level wave moves along or even south of the surface front. In those cases, there can be a 200 mb thick layer (or larger) where the best Q-vector convergence and frontogenesis are coincident. This can extend across a couple hundred miles. In those cases, one needs to examine a cross-section of frontogenesis and EPV to see where the frontal layer is nearly coincident with the least stable layer. This is best done prior to the development of precipitation in the model, which can impact the location of both frontogenesis and instability due to diabatic effects. On the right side of the cross-section, which is farthest south, the unstable layer is approximately 150 mb above the frontogenetic layer. However, toward the middle of the cross-section, the difference is less than 100 mb. This would suggest that 700 mb or 650 mb may be the best level to examine the frontogenesis and then look at EPV from 650–550 mb. This is because the decreased stability above the frontal surface would allow the frontal ascent to narrow and accelerate compared to farther south.

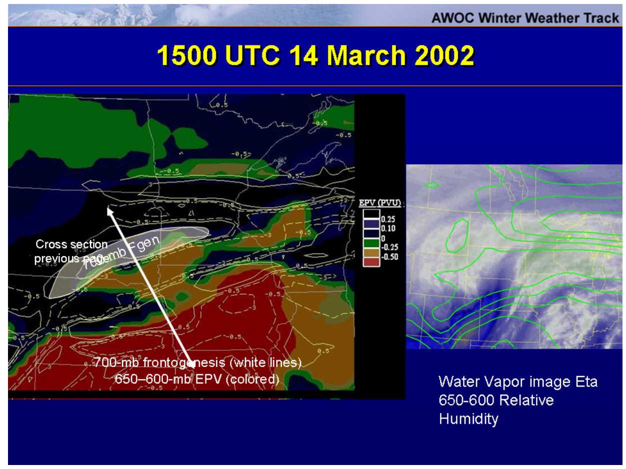
Student Notes:



27. 1500 UTC 14 March 2002

Instructor Notes: Once you identify the level(s) to examine frontogenesis and the layer to examine EPV, you can display a horizontal plot with both overlayed in order to see how both evolve over time. The first graphic shows 700 mb frontogenesis in the white lines and EPV_g in the 650 to 600 mb layer in color. Dashed white lines represent frontolysis and solid white lines represent frontogenesis. One can also examine the F_n divergence as discussed in Lesson 2. So we have also displayed 650 mb F_n divergence in the white lines and EPV, both from 1500 UTC 14 March. The convergence of F_n can be associated with upward vertical motion so either frontogenesis or convergence of F_n can be used to examine where the lift due to frontogenetic forcing will be located. When we look at frontogenesis, we are examining the ascent above the level of frontogenesis. Therefore, we examine EPV in the 650 to 600 mb layer which is where the ascending branch of the frontal circulation would be located. If F_n convergence is used, we are assuming that ascent is occurring at that level (650 mb) and so we examine EPV in a layer that includes the level we are displaying F_n convergence. Notice that in this case the 650 F_n convergence is co-located with the maximum of 700-mb frontogenesis. The diagnosis is the same in both figures – it is symmetrically unstable above the frontal boundary which would mean the potential an intense and narrow frontal band developing. Notice is it also very unstable across southern Nebraska and southern Iowa. However, as seen in the water vapor image, this is an area where the relative humidity is below 80 percent and where the dry slot is located. In fact, in many cases with frontal bands, the snow band will be at the northern edge of the dry slot where there is symmetric (and sometime conditional) instability and moisture. This region is not the most unstable area but it is the area where moisture, lift and instability combine to produce a heavy snowband.

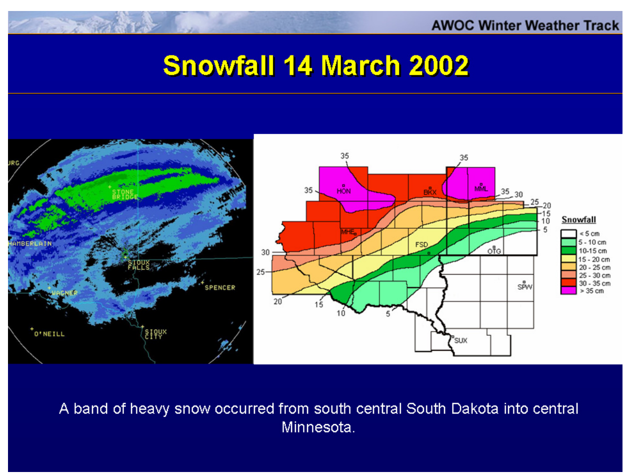
Student Notes:



28. Snowfall 14 March 2002

Instructor Notes: On the left is a radar picture which shows a band of heavy snow from the morning of 14 March. Snowfall in excess of one inch per hour was occurring at this time. The result was snowfall in excess of one foot (30 cm) across east central South Dakota and southwest Minnesota. Note the large gradient in snowfall between MML (Marshall, MN) and OTG (Worthington MN). These two cities are only separated by 60 miles but snowfall varied by over 12 inches. In this case the co-location of strong frontogenesis, low stability (or even instability) and a strong upper-level wave set the conditions for intense narrow band of snowfall.

Student Notes:



29. Interactive Quiz #2

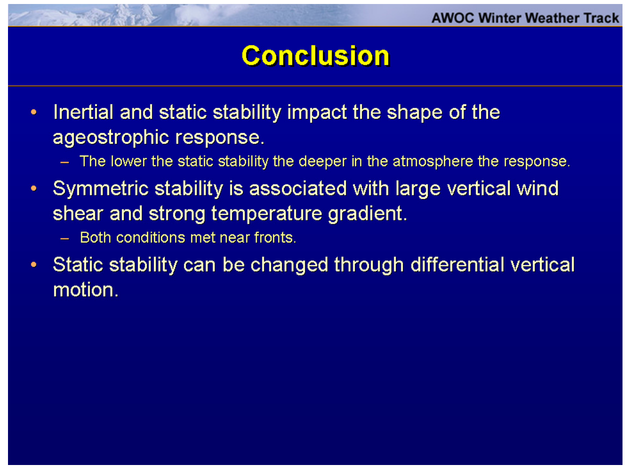
Instructor Notes: Take a few moments to complete this quiz.

Student Notes:

30. Conclusion

Instructor Notes: This is the first slide that discusses all the concepts covered in this presentation.

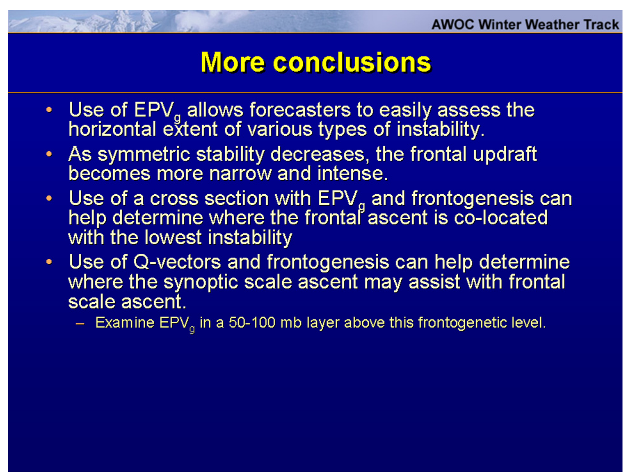
Student Notes:



31. More Conclusions

Instructor Notes: This is the second slide that summarized the content of this lesson.

Student Notes:



32. Have any Questions????

Instructor Notes: If you have any questions about this lesson, first ask your local AWOC facilitator. If you need additional help, send an E-mail to the address provided. When we answer, we will CC your local facilitator and may consider your question for our FAQ page. We strongly recommend that you take the exam as soon as possible after completing this lesson.

Student Notes: